1

The role of the Indian Ocean sector for prediction of the coupled Indo-Pacific system: Impact of atmospheric coupling

E.C. Hackert[1], A.J. Busalacchi[2,3], J. Carton[3], R. Murtugudde[1,3], P. Arkin[1], and M. N. Evans[1,4]

[1]Earth System Science Interdisciplinary Center, University of Maryland, College Park, Maryland, USA.

[2]University Corporation for Atmospheric Research, Boulder, Colorado, USA.

[3]Department of Atmospheric and Oceanic Sciences, University of Maryland, College Park, Maryland, USA.

[4]Department of Geology, University of Maryland, College Park, Maryland, USA.

Corresponding author: Eric Hackert (ehackert@essic.umd.edu)

Key Points:

- Indian Ocean teleconnections generate off-equatorial easterly winds and curl that act to amplify the oceanic Rossby waves in the Pacific

- These Rossby waves eventually positively impact the eastern Pacific via reflected western boundary then equatorial Kelvin waves

- Coupled hindcasts that include interannual forcing in the Indian Ocean significantly improve ENSO prediction skill from 3-9 months

12/13/2016

Abstract

22

23   Indian Ocean (IO) dynamics impact ENSO predictability by influencing wind and

24   precipitation anomalies in the Pacific.   To test if the upstream influence of the IO improves

25   ENSO validation statistics, a combination of forced ocean, atmosphere, and coupled models are

26   utilized.   In one experiment, the full tropical Indo-Pacific region atmosphere is forced by

27   observed interannual SST anomalies. In the other, the IO is forced by climatological SST.

28   Differences between these two forced atmospheric model experiments spotlight a much richer

29   wind response pattern in the Pacific than previous studies that used idealized forcing and simple

30   linear atmospheric models. Weak westerlies are found near the equator similar to earlier

31   literature. However, at initialization strong easterlies between 30^oS to 10^oS and 0^oN to 25^oN and

32   equatorial convergence of the meridional winds across the entire Pacific are unique findings

33   from this paper. The large-scale equatorial divergence west of the dateline and northeasterly-to-

34   northwesterly cross-equatorial flow converging on the equator east of the dateline in the Pacific

35   are generated from interannual IO SST coupling. In addition, off-equatorial downwelling curl

36   impacts large-scale oceanic waves (i.e. Rossby waves reflect as western boundary Kelvin

37   waves). After 3 months, these downwelling equatorial Kelvin waves propagate across the

38   Pacific and strengthen the NINO3 SST. Eventually Bjerknes feedbacks take hold in the eastern

39   Pacific which allows this warm anomaly to grow. Coupled forecasts for NINO3 SST anomalies

40   for 1993-2014 demonstrate that including interannual IO forcing significantly improves

41   predictions for 3-9 month lead times.

1. Introduction

El Niño/Southern Oscillation (ENSO) is the single most important and societally impactful mode of global climate variability on interannual time scales (e.g. [*Lau and Nath*, 2003], [*Glantz*, 2001], [*Horel and Wallace*, 1981]) but its prediction still has much to improve upon (e.g. [*National Academies of Sciences and Medicine*, 2016], [*National Research Council*, 2010]). For example, although six month lead forecasts from June 2014 confidently predicted strong ENSO warm phase conditions based on heat storage anomalies, an event did not develop, with the most likely explanation the failure to predict the absence of coupling between the ocean and the atmosphere [*McPhaden*, 2015].

There are several possibilities for this prediction failure which include 1) initial triggering events (i.e. westerly wind bursts) were out of sync with an amplifying mode and with the typical El Niño development timing (occurring 1 month earlier than for 1997 event) [*Menkes et al.*, 2014], 2) negative feedbacks such as upwelling ocean waves may have damped warm ENSO sea surface temperature anomalies (e.g. as happened for the 2002 event [*Hackert et al.*, 2007]), 3) stronger trade wind easterlies associated with the cool phase of the Pacific Decadal Oscillation (PDO) could have inhibited migration of the precipitation from the warm pool eastward [*Min et al.*, 2015], and 4) Indian Ocean (IO) dynamics may have anchored deep convection over the Indo-Pacific warm pool rather than allowing it to anomalously develop and couple with central/eastern Pacific SST anomalies [*Santoso et al.*, 2012].

Here we focus on the last possibility by isolating the impact of the IO atmosphere on the development of ENSO events. In previous work, [*Wu and Kirtman*, 2004], [*Annamalai et al.*, 2005], and [*Annamalai et al.*, 2010] proposed that cold IO sea surface temperature anomalies (SSTA) could generate an atmospheric Kelvin wave manifesting as equatorial westerly wind anomalies over the western Pacific, deepening the thermocline in the eastern Pacific via large-

66 scale oceanic Kelvin wave processes (e.g. [*Kessler et al.*, 1995]), and enhancing an ongoing El

67 Niño.

68   [*Wu and Kirtman*, 2004] compared coupled and decoupled general circulation model

69 (CGCM) experiments to test the impact of the IO on ENSO development. They showed that the

70 ENSO variability is reduced by a factor of two when the IO is decoupled (0.56^{o}C to 0.27^{o}C).

71 Composite equatorially averaged SST anomalies over longitude and time for the Pacific show

72 that maximum coupled SST anomalies at 165°W are located 15° east of those observed in the

73 uncoupled experiment (as in [*Yu et al.*, 2002]) and eastern upwelling associated with Kelvin

74 wave arrival is delayed by one month. The dominant period of variability is extended by half a

75 year by decoupling the IO, inconsistent with earlier work of [*Yu et al.*, 2002]. Using a linear

76 atmospheric model and idealized SSTA forcing, [*Wu and Kirtman*, 2004] found that the

77 influence of the IO is via the convective heating and modulation of the Walker circulation. IO

78 SST can induce anomalous Walker circulation over the eastern equatorial IO (EEIO) and

79 western-central Pacific through anomalous heating over the IO. For example, cold IO SST

80 induces development of anomalous westerlies to the east of Indonesia and easterlies to the west,

81 producing the atmospheric Kelvin wave pattern and a weakening of the Walker circulation [*Wu*

82 *and Kirtman*, 2004]. Therefore, IO SST forcing results indicate a mechanism that can enhance

83 westerlies over the western Pacific resulting in development of stronger El Niño conditions.

84   Consistent with these results, [*Annamalai et al.*, 2010] find that El Niño is much stronger

85 when occurring with Indian Ocean Dipole Zonal Mode (IODZM) and cold SST over the EIO and

86 in the Indonesian Seas (110°E-140°E, 10°S-0°). Winds are westerly over the Pacific from 130°E

87 east to 120°W and easterly over Indonesia west to the central IO indicating a weaker Walker

88 circulation. The eastern Pacific thermocline deepens and SST warms east of the dateline.

89   Precipitation is decreased from the central IO to 160°E in the Pacific and increased just east of

90   160°E and in the western Pacific. Further experiments with the Linear Baroclinic Model (LBM)

91   of [*Watanabe and Jin*, 2003] shows that EIO SST forcing generates an atmospheric Rossby

92   wave signature in the eastern IO but no signal in the Pacific, similar to the idealized forcing of

93   IODZM in [*Annamalai et al.*, 2005]. However the impact of the Indonesian Seas (120°E-160°E,

94   5°S-5°N) on the atmospheric linear model response is surprisingly strong and far reaching. Cold

95   SST forcing in the Indonesian Seas results in strong negative precipitation anomalies

96   ([*Annamalai et al.*, 2010] Figure 7b) creating an atmospheric Kelvin response that is evident by

97   easterlies over Indonesia and strong westerlies stretching to the east over the western equatorial

98   Pacific.

99   In summary, the IO can have strong influence for ENSO via atmospheric pathways.

100   Regional coupling of basin-scale IO and Indonesian Seas SSTA force enhanced westerlies over

101   the entire Pacific and increased ENSO forcing. The proposed mechanism is as follows: cold SST

102   over the EIO 1) reduces the normal east/west temperature gradient across the Pacific providing

103   favorable conditions for westerly anomalies to develop, 2) suppresses convection over the

104   Maritime continent and the resulting atmospheric Kelvin wave forces westerlies to the east, 3)

105   leads to convergence of the westerlies that enhance western Pacific precipitation, and 4) results

106   in an increased temperature gradient between Indonesia and the dateline further enhancing

107   westerlies. These westerlies then lead to downwelling oceanic Kelvin waves further enhancing

108   the growth of El Niño. Although [*Annamalai et al.*, 2010] conclude that regional IO SST and

109   heating anomalies are not the primary cause, but rather serve to enhance the development of El

110   Niño, the IO can have strong influences for ENSO via atmospheric teleconnections.

111   Unfortunately, all previous studies (i.e. [*Wu and Kirtman*, 2004], [*Annamalai et al.*, 2005], and

112   [*Annamalai et al.*, 2010], etc.) used idealized SSTA patterns and simplified linear atmospheric

113   models to show impacts of the IO on the wind field without assessing observed ENSO

114   predictability.

115   Our approach is to use a combination of ocean-only and coupled models to diagnose the

116   impact of the IO atmospheric teleconnections to ENSO predictability. We will use similar

117   techniques as in previous studies ([*Yu et al.*, 2002], [*Wu and Kirtman*, 2004], [*Annamalai et al.*,

118   2005], [*Annamalai et al.*, 2010], and [*Santoso et al.*, 2012]) to diagnose the impact of the

119   interannual SSTA forcing in the IO. Namely, we difference results from experiments that fully

120   couple the SSTA throughout the Indo-Pacific region with ones that decouple the IO by forcing

121   with climatological SST in this region. Unlike all previous studies, however, we develop

122   realistic coupled hindcast experiments using realistic interannual anomalies as forcing, and

123   validate the results against observations of forecasted ENSO state. The overriding hypothesis

124   that we wish to test is that the upstream influence of the Indian Ocean improves ENSO

125   predictions through mechanisms associated with the atmospheric bridge as coined by [*Alexander*

126   *et al.*, 2002]. We will use "atmospheric bridge", "atmospheric teleconnection" or "atmospheric

127   impact" interchangeably throughout the following text for this impact.

128   The organization of this paper is as follows. Section 2 covers ocean and coupled models

129   and Section 3 describes the simulations and analysis techniques. Section 4 reports the various

130   model results, Section 5 contains the discussion, and Section 6 summarizes the paper results and

131   provides conclusions.

132

2. Models

2.1 Ocean Model

The ocean general circulation model (OGCM) that is used in this study is the primitive-equation, sigma-coordinate model with variable depth oceanic mixed layer of *Gent and Cane* [1989]. It is described and validated in a series of simulation studies of circulation in all three tropical ocean basins [*Hackert et al.*, 2001; *Murtugudde and Busalacchi*, 1998; *Murtugudde et al.*, 1996; *Murtugudde et al.*, 1998]. Solar radiation (Earth Radiation Budget Experiment - ERBE) and interannual precipitation from the Global Precipitation Climate Project - GPCP [*Adler et al.*, 2003] are specified externally. Monthly anomalies of the cloud data [NCEP Reanalysis *Kalnay et al.*, 1996] are added to the Interannual Satellite Cloud Climatology Project –ISCCP annual cycle [*Rossow and Schiffer*, 1991] in order to provide a more realistic mean.

Our OGCM uses the hybrid vertical mixing scheme of *Chen et al.* [1994] which combines the advantages of the traditional bulk mixed layer of *Kraus and Turner* [1967] with the dynamic instability model of *Price et al.* [1986]. This allows simulation of all three major processes of oceanic vertical turbulent mixing - atmospheric forcing is related to mixed layer entrainment/detrainment, gradient Richardson number accounts for shear flow, and instantaneous adjustment simulates high frequency convection in the thermocline. The vertical structure consists of a variable depth mixed layer and 19 sigma layers with a deep motionless boundary being specified as $T_{bottom} = 6^{\circ}$C and $S_{bottom} = 35$ PSU.

The ocean model configuration used for all simulations covers the tropical Indo-Pacific basin (34°E -76°W, 30°S-30°N) with a homogeneous 1° longitudinal grid and a variable latitudinal grid (down to $1/3^{\circ}$ within 10° of the equator). This resolution is dense enough to allow mesoscale eddies and realistic flow. Surface fluxes are calculated interactively by coupling the OGCM to a thermodynamic atmospheric mixed layer model [*Murtugudde et al.*, 1996] thus

157 allowing feedbacks between SST, SSS, and surface fluxes. The open boundaries are treated as a

158 sponge layer within $10^°$ of the north and south borders smoothly relaxing to World Ocean Atlas

159 2009 (WOA09 - [*Antonov et al.*, 2010]. Note that this model surface is allowed to vary freely as

160 a natural boundary condition ([*Huang*, 1993]) and only relaxes back to Levitus temperature and

161 salinity ([*Locarnini et al.*, 2010]) within the north and south boundary sponge layers.

162   The model is spun up from rest using climatological winds with the initial conditions

163 derived from WOA09 data and is allowed to come to equilibrium after 30 years of forcing by the

164 *ECMWF* [1994] analysis climatology.   Interannual runs are initialized from this climatological

165 spin-up starting in 1975 and the wind speeds required for sensible and latent heat fluxes are

166 computed from interannual ECMWF 10 m wind converted to stress using the bulk formula ($\rho =$

167 1.2 kg/m^3, C$_D$ = 1.2x10^3).

168 **2.2 Atmospheric Model**

169   An intermediate complexity atmospheric general circulation model (AGCM), the

170 International Centre for Theoretical Physics AGCM (nicknamed SPEEDY, for "Simplified

171 Parameterizations, primitivE-Equation DYnamics" - [*Molteni*, 2003]; [*Kucharski et al.*, 2006])

172 provides an accurate atmospheric model response, yet is highly computationally efficient. We

173 use SPEEDY Version 41, which has global T30 resolution (roughly 3.75^o) with 8 standard sigma

174 layers (925 – 30 mb) and surface information. The winds in the tropics have been improved by

175 adding cumulus momentum transport (CMT) to the convective parameterization code using the

176 technique of [*Kim et al.*, 2008].   This technique transports momentum downward within

177 subsidence regions surrounding regions of convection.   Adding CMT to the atmospheric model

178 shifts wind and western Pacific precipitation anomalies eastward, which are more in line with

179 observations.   In addition, the meridional extent of the wind anomalies is expanded due to the

180 incorporation of the CMT. For example, [*Kim et al.*, 2008] show that 850 mb westerlies during

181 ENSO expand from 15°S-0°N without CMT to 15°S-10°N with CMT.   Implementation of CMT

182 has also shown to improve intraseasonal precipitation, SST and winds such as those associated

183 with Madden Julian Oscillations [*Zhou et al.*, 2012]. Hence, effects of the IO on the coupled

184 ENSO system should be distinguishable within SPEEDY simulations via analysis of wind and

185 precipitation anomalies on intraseasonal to interannual timescales.

186   The winds and precipitation from SPEEDY have similar validation statistics as for other

187 atmospheric models and observations. For example, the mean 925 mb winds of SPEEDY over

188 the tropics closely match the European Centre for Medium-Range Weather Forecasts reanalysis

189 (ERA - [*Gibson et al.*, 1997]). For precipitation, all the major features of the observations

190 (CMAP from [*Xie and Arkin*, 1998]) are reproduced by SPEEDY. However, the SPCZ has less

191 abundant rainfall and relatively more precipitation over the tropical IO [*Molteni*, 2003] and

192 SPEEDY tends to underestimate the zonal wind anomalies associated with ENSO events (see

193 Figure 9 of [*Kroeger and Kucharski*, 2011]). For a full description of latest version of SPEEDY

194 see [*Kucharski et al.*, 2013].

195 **2.3 Coupled Model**

196   Coupling of an intermediate complexity atmospheric model such as SPEEDY is justified

197 since the atmospheric time scale is much shorter than that of the ocean. The SPEEDY AGCM

198 has been successfully coupled with other ocean models for the Pacific (e.g. [*Kucharski et al.*,

199 2011]), Indian ([*Kucharski et al.*, 2006]), and Indo-Pacific regions ([*Bracco et al.*, 2005]).

200 Similar to the implementation of [*Kroeger and Kucharski*, 2011], we use the technique of

201 anomaly coupling to couple the ocean and atmospheric models. Within the tropical Indo-Pacific

202 region, our ocean model SSTA forces the SPEEDY AGCM. For the rest of the globe, the

203   observed SSTA of HadISST [*Rayner et al.*, 2003] is used and climatological values are used for

204   other atmospheric model boundary conditions such as surface albedo, climatological SST, sea

205   ice, snow depth, vegetation, heat flux parameters, and soil moisture (matching those described in

206   [*Kucharski et al.*, 2013]). The SST anomaly is formulated with respect to the ECMWF/GPCP

207   climate experiment (described in Section 2.1). The atmospheric model is then spun up for 1

208   month using this SSTA and anomalies of surface zonal and meridional wind stress (τ'_x, τ'_y) and

209   precipitation (P') are formulated with respect to the mean seasonal cycle over 1993-2014 of the

210   similarly forced atmospheric model. These anomalies are subsequently added back to the

211   ECWMF and GPCP climatologies to force the next month of the ocean model simulation. While

212   more efficient than the operational coupled models which use high resolution and more complete

213   atmospheric models, the coupling of SPEEDY nevertheless allows adequate physics to quantify

214   the impact of the atmospheric bridge on ENSO prediction. Using this technique, one year

215   forecasts are completed for each month for 1993-2014.

216   In order to validate the coupled model, NINO3 SST anomalies are compared against

217   observations. SPEEDY coupled model has significantly better correlation (Figure 1a) and

218   amplitude validation (Figure 1b) than observational persistence after 4 months and 3.5 months,

219   respectively. The correlation of SPEEDY versus observations remains significant at $p<0.05$

220   (df=35) for 8 month lead times. As a forecast skill comparator, we include the Climate Forecast

221   System Reanalysis Reforecast ([*Saha et al.*, 2014]) as a reference (Figure 1 - black line). These

222   coupled hindcasts are comprised of the atmospheric assimilation/model with resolution ~38 km

223   (detailed in [*Saha et al.*, 2010]) along with the MOM4 ocean model ([*Griffies et al.*, 2004]) with

224   0.5^o resolution within 30^oN-30^oS and the Global Ocean Data Assimilation System (GODAS)

225   ocean assimilation ([*Behringer*, 2007]) of all available oceanic in situ data. The CFSRR model

226   was chosen to substantiate our coupled model results since it is a well-known, state of the art,

227   operational coupled model (i.e. the reanalysis, reforecast version of CFSv2).   For these long

228   validation runs (i.e. January 1993 to March 2011), correlation between forecasted and observed

229   NINO3 SSTA are equivalent (Figure 1a). For validation by root mean square difference

230   (RMSD), results from our coupled model outperform the CFSRR for all lead times. CFSRR

231   RMSD errors reach 1.4^{o}C at 10 month lead times, whereas SPEEDY coupled model RMSD

232   errors are 1°C for the same lead time. Based on these diagnostics, our coupled model validates at

233   least as well as the NOAA operational model.

3. Simulations and Analysis

3.1 Uncoupled Simulations

SSTA is calculated over 1993-2014 from the ocean model that is forced using all available observed winds, cloudiness, and precipitation (described in Section 2.1). SPEEDY atmosphere-only experiments are then initiated using different ocean model SSTA for the IO basin to isolate the ocean forcing impacts via the atmosphere on the coupled Indo-Pacific system. In this work, the Pacific (abbreviated PAC) is defined as 30^oN-30^oS, 130^oE-70^oW and the Indian Ocean (IO) is defined as 30^oN-30^oS, 30^oE-129^oE. Outside the tropical Indo-Pacific region, the interannual SSTA from the Hadley Centre (HadISST, [*Rayner et al.*, 2003]) is used to force the global SPEEDY model. Within the Indo-Pacific region, experiments were initiated that are designed to isolate the impact of the IO region surface forcing on the atmosphere.

Table 1 shows the complete set of experiments performed for this study. The experiments either use interannual (i.e. INT) SSTA forcing or climatology seasonal cycle (CLIM) SST separated by basin, PAC and IO. For example, forcing SPEEDY using interannual SSTA for the Pacific and IO is abbreviated as INT_PAC, INT_IO. Following the similar methodology of e.g. [*Wu and Kirtman*, 2004], we subtract the results from these different experiments in order to isolate the impact of the IO sector ocean forcing. Thus, subtracting INT_PAC, CLIM_IO results from INT_PAC, INT_IO will isolate the impact of the IO SSTA via the atmospheric teleconnections to the Pacific.

3.2 Coupled Simulations

A series of coupled experiments designed to isolate the full impact of the interannual IO SSTA forcing is executed. Operationally the same initial conditions are used as for the uncoupled simulations, since the goal is to completely eliminate any potential impacts caused by

258   different initialization. However, within the anomaly coupling procedure we replace the regional

259   SST anomaly with zeros effectively substituting the forecast interannual SST forcing with

260   climatological seasonal cycle values. A series of two 12 month coupled experiments were

261   completed for each month from 1993-2014: for each experiment, a total of 264 12 month runs

262   were completed. The two experiments are 1) INT_PAC, INT_IO (interannual SST anomaly

263   forcing for the Pacific and the IO) and 2) interannual Pacific, climatological IO (INT_PAC,

264   CLIM_IO). As in the uncoupled experiments (section 3.1), the experiments with similar Pacific

265   forcing are compared, e.g. INT_PAC, INT_IO versus INT_PAC, CLIM_IO.

266   **3.3 Analysis Techniques**

267   Unlike all previous similar research, the impact of the IO on ENSO predictability is

268   evaluated using observed quantities. Specifically, the observed NINO3 (5^oS-5^oN, 90^oW-150^oW)

269   SSTA from [*Reynolds et al.*, 2002] is used to validate all coupled and forced model results using

270   correlation and RMSD. For correlation, the effective degrees of freedom (df) is calculated using

271   the technique of [*Quenouille*, 1952] (pp. 168-170) with the equation:

272   df=N/(1.+2.*r_a(1)*r_b(1)+2.*r_a(2)*r_b(2)+2.*r_a(3)*r_b(3)) where N is original number of

273   observations and r_a and r_b are autocorrelations for time series a and b, respectively for 1, 2, and 3

274   months lags (indicated by indices).   After the effective degrees of freedom are calculated the

275   Students T test is used to establish significance of correlation values. For all statistics, a

276   probability of less than p=0.05 that a correlation is zero will be considered statistically

277   significant and interpretable.   In order to test the impact of the IO atmospheric teleconnections

278   to ENSO, forecast lead time correlations will be compared between different coupled model

279   results.   Since all experiments are validated against observed SST anomalies, and so share a

280   common variable, these correlations are not independent (known as correlated correlations).

281 Therefore, the Steiger's Z-test [*Steiger*, 1980] will be utilized to test the significance of the

282 differences between correlations as applied in [*Uehara et al.*, 2014].   The details follow:

283 $Z = [Z_{aa} - Z_{ba}] * \dfrac{\sqrt{N-3}}{\sqrt{2}*[1-r_{aa}]*h}$ where Z_{ao}, Z_{bo} are Fisher Z transformations of r_{ao} and r_{bo} (the

284 correlation of experiments a and b, respectively versus observations (o), N is the number of

285 observations and r_{ab} is the correlation between the two forecast experiments, a and b, $h =$

286 $\dfrac{1-(f*rr^2)}{1-rr^2}, f = \dfrac{\frac{1-r_{aa}}{a}}{2*[1-rr^2]}$, and $rr^2 = \dfrac{r_{ao}+r_{bo}}{2}$. This technique has the benefit of producing a

287 statistic that is normally distributed, so for $|Z| > 1.96$, $p <= 0.05$. In addition, the lead time

288 amplitude of the various forecasts will be validated against observed NINO3 SST anomaly using

289 RMSD. Forecast standard deviation and mean will also be used to compare different forecasts

290 versus observed values.   The forecast mean (standard deviation) is simply the mean (standard

291 deviation) across all forecasts, on lead times from 0 to 12 months.

292   In order to quantify the impact of Kelvin and Rossby wave propagation on El Niño

293 events, sea level anomalies are first converted to geostrophic currents using the methodology of

294 [*Picaut and Tournier*, 1991]. Next, the technique derived by [*Delcroix et al.*, 1994] is used to

295 separate the sea level anomalies geostrophic current data into Kelvin and Rossby components.

296

4. Results

4.1 Forced Ocean and Atmospheric Model Results

The results of the SPEEDY atmospheric model differences described in Section 3.1 are designed to isolate the impact of the IO and are presented in Figure 2a-f for zonal and meridional wind stresses, precipitation, curl and divergence of the wind stress, respectively. By differencing experiments with full coupling in the IO minus those with decoupled IO, the impact of the variations in the IO summer monsoon is readily apparent for both precipitation and wind stress. For precipitation (Figure 2c), positive anomalies can be seen stretching from the equator to 10^oN in the eastern Arabian Sea (AS), at 7^oN to 12^oN in the Bay of Bengal (BOB), and 10^oN to 20^oN in the South China Sea (SCS). Abundant rainfall is consistent with convergence of the monsoon flow starting south of the equator as southeasterlies, recurving to southerlies near the equator and decelerating as southwesterlies in the AS, BOB, and SCS (Figure 2d and f). These southwesterlies converge into northeasterlies found north of 10^oN in the BOB and SCS. In the Southern Hemisphere, between 10^oS and the equator, negative precipitation anomalies are generally found west of 95^oE. These precipitation patterns are consistent with the general divergence of the winds in these regions as they feed into the northward monsoon flow. In addition, positive precipitation is found over the southern Indonesian islands between 95^oE and 130^oE stretching between 12^oS to 5^oS. This feature is due to the onshore convergence of the westerlies found west of 95^oE and is consistent with anomalies that are associated with the transition of the northwest to summer monsoons in boreal spring. In the southwest IO between 25^oS-15^oS and west of 80^oE, a band of positive precipitation is evident just to the east of Madagascar and is the result of southeasterlies to the south converging with northwesterlies to the north (Figure 2f). Warm SST leads to atmospheric advection, enhanced inflow, convergence and abundant atmospheric convection and precipitation. The patterns in the IO are consistent

321 with the anomalies of the monsoon wind and precipitation patterns (those patterns associated

322 with interannual minus climatological seasonal cycle SST forcing in the IO).

323 In the Pacific, generally positive precipitation anomalies are found off the equator at 5°N,

324 15°S centered near 160°W (Figure 2c). In addition, positive anomalies are seen in the upwelling

325 region of the eastern equatorial Pacific east of 130°W and in the South Pacific Convergence

326 Zone (SPCZ) at 10°S, 160°E. North of 10°N and south of 20°S, negative precipitation anomalies

327 are simulated. Strong easterly anomalies can be seen between equator and 20°N and south of

328 15°S across the entire Pacific basin (Figure 2a). North of 20°N, strong westerly anomalies

329 prevail. Between 10°S and the equator, wind differences are generally very weak but westerly.

330 The meridional winds (Figure 2b) converge to roughly 5°S with northerlies to the north and

331 southerlies to south especially east of the dateline. Along 20°N winds are generally divergent for

332 the meridional wind plot.

4.2 Coupled Model Results

333

334 The validation of the two coupled simulations versus observed NINO3 SSTA over all 12

335 month lead times, 1993-2014, is presented in Figure 3a. Both simulations are significantly

336 correlated with observations out to 8 and 5.8 months for INT_PAC, INT_IO and INT_PAC,

337 CLIM_IO, respectively. After 3 month lead times, the correlation of the full coupling begins to

338 outperform the INT_PAC, CLIM_IO coupled simulation. Correlation differences climb to

339 r=0.16 by 7 month lead times. At this time the Steiger Z Test ([*Steiger*, 1980]) shows that the

340 differences are significant (thick dashed line on top x axis in Figure 3a). After that, the

341 differences drop to about r=0.1 out to 10 month lead times when the Figure 3 differences are no

342 longer significant. The important result of this plot is that the interannual forcing of the IO

343 significantly improves coupled forecasts for ENSO. Results for RMSD of simulated versus

344　observed NINO3 SSTA (Figure 3b) are consistent with results for correlation (Figure 3a) and

345　quantify the validation error amplitude.

346

5. Discussion

Over most of the forecast period and particularly between 3 to 9 month lead times, interannual SST forcing in the IO improves correlation and reduces the RMS differences between observed NINO3 SSTA and those simulated by the fully coupled model. Therefore, these results suggest that including the impacts of the IO atmospheric teleconnection serves to improve the coupled predictability as validated with real observations over 1993 to 2014.

The next step is to examine why the interannual SST forcing of the IO improves the coupled forecasts. Figure 4 shows the mean and standard deviation of the NINO3 SSTA of all the 12 month forecasts from 1993-2014. The mean plot (Figure 4a) shows that the experiment with interannual IO forcing has higher mean values (relative warming signal in the NINO3 region) after 3 months. On the other hand, the INT_PAC, INT_IO and INT_PAC, CLIM_IO standard deviation lines in Figure 4b practically overlay one another. This result is different than that of [*Santoso et al.*, 2012] and [*Wu and Kirtman*, 2004] who found that the IO increased the amplitude of ENSO events. On the contrary, we conclude that the interannual signal in the IO serves to warm the mean state in the eastern Pacific after 3 months rather than impact the variability.

To further diagnose the source of the warming after 3 months, the mean forecast difference, INT_PAC, INT_IO minus INT_PAC, CLIM_IO, is presented for all 12 months of lead times. Equatorial longitude versus time plots track the evolution over the average forecast in Figure 5. Early in the average forecast difference, prior to month 3, easterly winds along the equator between 140°E-160°E and between 180°-140°W (Figure 5c) set off upwelling Kelvin waves (Figure 6a), cooling SSTA (Figure 5a), and inducing westward flow across the entire Pacific (Figure 5d). This is consistent with the general upwelling favorable curl in the initial conditions (Figure 2e) between 15°S-10°N. After this slight upwelling and cooling in the central

371 Pacific associated with equatorial easterlies, the SST in the NINO3 region begins to warm after 3

372 month lead times (Figure 5a).   In the east, westerlies centered at approximately 130°W generate

373 a downwelling Kelvin wave that arrives at the eastern boundary at month 4 (Figure 6a). At this

374 time the NINO3 region begins to warm (Figure 5a). In the west, westerlies on the equator west

375 of 140°E and near the dateline act in the equatorial Pacific setting off a second downwelling

376 Kelvin wave, which is identified by positive sea level anomaly and eastward flow (Figure 5b, d

377 and Figure 6a) that starts in month 4 and traverses the Pacific and arrives at the eastern boundary

378 by month 6. The cumulative effects of the downwelling Kelvin waves after month 5 are to

379 continue to warm the NINO3 region. For month 6 through 8, warmest SSTA is building in the

380 central Pacific between 160°E and 140°W. Westerlies to the west and easterlies to the east

381 converge into this warm region (Figure 5c) near the dateline. A weak upwelling Kelvin wave

382 (Figure 6a dashed line) is initiated that is associated with these easterlies east of the dateline. At

383 this time, the prevailing eastward flow is interrupted and westward currents are found in the

384 eastern Pacific between months 6-8 (Figure 5d). The SSTA in the NINO3 region briefly cools

385 (Figure 5a) in month 7. At this same time (forecast months 6-8), westerlies prevail from the

386 western boundary all the way to the dateline. The next downwelling Kelvin wave is initiated in

387 the west and arrives at the eastern boundary roughly at forecast month 9. As it enters the NINO3

388 region this downwelling Kelvin wave warms the SSTA. The warmest SSTA fills in to the west

389 and by month 10 some of the warmest SSTA is located just east of the dateline. Bjerknes

390 feedback becomes entrenched and westerlies to the west of the SSTA maximum converge with

391 easterlies to the east (Figure 5c).

392   Careful examination of the equatorial signal and Kelvin/Rossby wave decomposition of

393 the ocean waves helps to explain the timing and sign of the differences in the mean state in

394 Figure 4a. After 3 month forecasts, downwelling Kelvin waves that are forced by westerlies in

395 the western Pacific, start to warm the eastern Pacific. After 7 month lead times, the Bjerknes

396 feedback mechanism begins to lock in leading to enhanced westerlies over the western Pacific

397 (Figure 5c) and growth of the air-sea coupled mode. The atmospheric impact of including the

398 interannual forcing in the IO is to impart a large-scale downwelling favorable signal in the

399 Pacific, increasing the warming in the NINO3 region after the 3 months. By 7 month lead times,

400 the El Niño signal is enhanced/reinforced due to Bjerknes feedback.

401 To a large part, the previous discussion reinforces the conjecture of [*Annamalai et al.*,

402 2010] who suggested that the impact of the IO would be to enhance the westerlies along the

403 equator and amplify an ongoing El Niño. However, examination of the various fields besides

404 equatorial Hovmöller plots suggests that the initialization and growth of the warming in the

405 NINO3 region is influenced by off-equatorial factors and by not only zonal but also meridional

406 wind stress. Therefore, the discussion will now focus on the Pacific basin using plots of the

407 mean forecast for 3, 5, 7, and 10 month lead times for SSTA, sea level anomaly, curl and

408 divergence differences for INT_PAC, INT_IO minus INT_PAC, CLIM_IO results (Figure 7 to

409 Figure 10, respectively).

410 On the equator, Ekman pumping velocity is undefined (since the Coriolis parameter is in

411 the denominator). However, near the equator the wind stress divergence can be diagnosed to

412 infer regions of upwelling or downwelling. By month 3 of the mean forecast, divergence

413 corresponds to upwelling between 140°E-150°W on the equator and convergence is found

414 between 140°W-110°W (Figure 7d). Off the equator, downwelling favorable curl (curl <0 in the

415 Northern Hemisphere and > 0 in the Southern Hemisphere) can be seen west of 140°W generally

416 within 10° of the equator (Figure 7c). West of 160°W in the far western Pacific downwelling

417 favorable curl off the equator corresponds to positive sea level anomalies off New Guinea

418 (Figure 7b). This feature is important since it initiates the transition from the upwelling prior to

419 month 3 across the basin to overall downwelling after that time. In other words, the off-

420 equatorial curl initializes a downwelling Rossby wave and positive sea level anomaly in the far

421 western Pacific that soon reflects as a downwelling Kelvin wave that begins the eventual

422 transition to warm SSTA in the NINO3 region by month 7.   This is an instance where the off-

423 equatorial signal (a downwelling Rossby wave) contributes to converting upwelling to

424 downwelling along the equator and so features prominently in ENSO predictability.

425 Unfortunately, this downwelling in the west is not well represented by the Kelvin/Rossby

426 decomposition plot (Figure 6) since it lies west of 160°E which is the western extent of the land-

427 free symmetric box that is required by this decomposition analysis. However, the subsequent

428 downwelling Kelvin wave (spawned from the reflected Rossby wave) in the far west starting in

429 month 4 is well diagnosed.

430 　　　To the east, a pair of upwelling-favorable (negative in the Northern Hemisphere positive

431 in the Southern Hemisphere) curl patches are located within 15°S-10°N between 140°W-110°W

432 (Figure 7c).   This feature corresponds with a pair of negative sea level anomalies centered at

433 5°N and 12°S at 130°W and is identified as an upwelling Rossby wave in the Kelvin/Rossby

434 diagnosis in Figure 6b. East of 160°W, an upwelling Rossby wave at 140°W acts to shoal the

435 thermocline at 5°N and 10°S reshaping the meridional gradient to help focus the downwelling

436 Kelvin wave train along the equator coming later in the average forecast.

437 　　　In summary, downwelling Rossby waves forced by wind stress curl off the equator in the

438 far western Pacific reflect to downwelling Kelvin waves eventually transitioning the NINO3

439 region to warming. The upwelling Rossby wave at 140°W at month 3 shapes the meridional

440 gradient to focus intensification on the equator.

441 By month 5 the downwelling Rossby wave hitting the western boundary in month 3 has

442 reflected into a downwelling Kelvin wave and this wave has propagated eastward across the

443 Pacific as far as ~140°W (Figure 8b and Figure 6a). The effects of this downwelling Kelvin

444 wave are demonstrated by positive sea level and SST anomaly throughout the waveguide (+/- 2°)

445 across the entire Pacific (Figure 8a). The upwelling features in Figure 8c are echoed in negative

446 sea level at 5°N and 10°S at 145°W (Figure 8b). The NINO3 region is warming and SSTA is

447 largest at about 120°W on the equator. It is also interesting to note that the warmest SSTA is just

448 south of the equator whereas the sea level anomaly maximum is centered on the equator.

449 Convergence found on the equator and positive curl and downwelling just to the south (5°S-0°S,

450 140°W-110°W) coincide with maximum SSTA. Going from 5°N to 12°S along 130°W winds

451 are starting northerly recurving to northwesterlies just south of the equator. West of the dateline

452 and south of 5°S, pervasive positive curl (downwelling favorable) is collocated with positive sea

453 level anomaly against New Guinea and Australia coasts. To the north of the equator, positive

454 curl and upwelling are found with negative sea level west of 150°E off the Philippines. Thus the

455 southeasterlies to the south recurving to southwesterlies north of the equator in the far western

456 Pacific act to deepen and shoal sea level, respectively. To reiterate, not only are the equatorial

457 signals important for the diagnosing the impact of the IO on ENSO, but the off-equatorial

458 impacts such as oceanic Rossby wave formation and propagation are also important.

459 By month 7 the second Kelvin wave has reflected at the eastern boundary as a

460 downwelling Rossby wave as evident by positive sea level at 10°N, 5°S at 120°W (Figure 9b and

461 Figure 6b). Another positive sea level and SSTA maximum is centered on the equator at

462 ~170°W (Figure 9a, b). Equatorial westerlies, best demonstrated by the westerly wind burst in

463 Figure 5c that extends from the western boundary to 160°E (note the 2×10^{-3} N/m^2 contour), force

464 this downwelling Kelvin wave. To the north, downwelling curl corresponds to positive sea level

465 and to the south negative sea level is collocated with upwelling favorable wind stress curl at

466 10°N, 10°S at ~150°W, respectively. Off the equator, west of 160°E the curl is positive to the

467 north and this forces upwelling 0-10°N and negative sea level. It is also interesting to note that

468 upwelling curl within 5° of the equator in the NINO3 region (particularly at 120°W) is causing

469 weak upwelling and cold SSTA at 5°N driving the warmest SST south of the equator (Figure 9a).

470 By month 10 the Bjerknes feedback has locked in (Figure 10). SSTA is positive

471 throughout the equatorial band between 160°E to the eastern boundary. The negative sea level

472 horseshoe pattern is evident off the equator in the west and positive values east of 160°E near the

473 equator typically associated with a mature El Niño. On the equator, winds are diverging between

474 150°E-175°E and converging to the east of there between 150°W-100°W. To the east of 150°W

475 and off the equator, the curl and sea level are in good agreement. At 10°N, 150°W, downwelling

476 curl corresponds with positive sea level anomaly. At 2°N, 135°W positive curl overlays with a

477 small region of negative sea level. Just south of the equator at 135°W, downwelling curl

478 coincides with a maximum of sea level and SSTA. Further to the south at 10°S and 140°W

479 upwelling curl and negative sea level coincide. Thus the pattern of upwelling/downwelling curl

480 of the wind maintains the meridional sea level gradient east of 160°W. Off the equator west of

481 160°W, upwelling curl is acting to reinforce the negative sea level off the Philippines and off the

482 coast of New Guinea and Australia. To summarize, the mean forecast by 10 month lead times

483 shows winds that are primarily diverging away from the equator west of the dateline

484 (southwesterlies to the north and northwesterlies to the south of the equator) and converging

485   towards the equator east of the dateline (with northeasterlies north of the equator slowing and

486   turning towards northwesterlies at the equator). These diagnostics of the average coupled

487   forecast reveals that the response of the ENSO system in the Pacific is more complicated than

488   simply triggering a westerly wind burst in the western Pacific on the equator and setting off

489   downwelling Kelvin waves eventually warming the NINO3 region as assumed by [*Annamalai et*

490   *al.*, 2010] and others. Our diagnosis suggests that the atmospheric response is more complicated

491   than previously thought, and a previously unaccounted-for significant signal corresponds to

492   strong easterlies south of 15°S and between 0°N to 20°N.

493   Our results may be usefully contrasted with previous work employing simpler

494   atmospheric models and idealized surface forcing. Similar to previous work, there are weak

495   westerlies near the equator (10°S-0°N) across the entire Pacific. These relatively weak winds we

496   simulate near the equator match those of [*Annamalai et al.*, 2005] who used a simple linear

497   atmospheric model (a moist linear baroclinic model – LBM of [*Watanabe and Jin*, 2003]) to

498   show that the atmospheric Kelvin wave of the western dipole of the IODZM cancels that of the

499   eastern dipole (their Figures 8b, 10d). However, our results show the importance of not only the

500   near-equatorial winds but also the off-equatorial zonal and meridional winds for the diagnosis of

501   the IO SSTA teleconnections to Pacific ENSO (Figure 2a).

502   The off-equatorial easterlies represented in Figure 2a are a prominent feature but are

503   lacking in previous studies (e.g. [*Wu and Kirtman*, 2004] Figure 7d, [*Annamalai et al.*, 2010]

504   Figure 7d). There are several reasons why the simple atmospheric models that were used to

505   highlight the IO atmospheric teleconnections to the Pacific ([*Wu and Kirtman*, 2004] and

506   [*Annamalai et al.*, 2010]) might lack the off-equatorial easterlies simulated in SPEEDY. In

507   previous results, the IO SST field is idealized in some way or another. Either the 1st EOF of the

508 coupled model results or the SST differences with and without IODZM for El Niño is utilized to

509 simplify the SST forcing. In addition, these models have been linearized about different mean

510 states for specific seasons (JJAS and MJ, respectively) so the seasonal cycle remaining in

511 SPEEDY may play a role in forming the off-equatorial easterlies. When we limit the SSTA

512 forcing in the IO to just the 1st EOF of the simulated SSTA (not shown but a similar pattern as

513 [*Wu and Kirtman*, 2004] Figure 7a for the IO) for the forced SPEEDY AGCM, the anomalous

514 westerlies near the equator are enhanced: the atmospheric Kelvin wave is present as in previous

515 research but the easterlies off the equator remain.

516   The lack of a strong signal off the equator for the LBM results of the previous authors is

517 surprising considering the results of [*Watanabe and Jin*, 2003]. They used a similar model

518 (LBM) as previous authors and forced it with El Niño minus La Niña observed SSTA limited to

519 the IO region. Their Figure 8c indicates that a basin-scale cooling in the IO (their M3 region)

520 results in a positive precipitation response symmetric within 15° of the equator that is centered at

521 140°E. West of the dateline, the 850 mb streamfunction response to this heating shows nearly

522 collocated cyclonic flow and easterlies between 10°N-35°N, westerlies for 5°N-10°S, and

523 easterlies between 15°S-30°S west of the dateline, which are broadly similar to our results.   On

524 the other hand, our precipitation results have this maximum centered to the east, at roughly

525 160°W (Figure 2c), so this displacement of the precipitation heating might explain the elongated

526 off-equatorial easterlies found in the SPEEDY results (Figure 2a).

527   Another potential difference between previously reported results and the current wind

528 results is the amplitude and location of the precipitation anomalies in the Pacific.   For example,

529 [*Annamalai et al.*, 2010] shows (their Figure 2b) the strong positive precipitation anomalies

530 centered at 10°S and 5°N centered at 180°, roughly similar to our results. However, for these

531   previous results there is also a strong negative anomaly with similar zonal extent and amplitude

532   to the west, centered at 150°E that may act to offset any off-equatorial signal in the winds.   Our

533   precipitation results (Figure 2c) show that there are no such offsetting precipitation anomalies to

534   the west of the main positive values located between 160°E to 160°W centered at 10°S and 5°N.

535   Therefore, off-equatorial easterlies are not opposed by westerlies for the SPEEDY results.

536   The last and most likely potential reason for off-equatorial easterlies may be the

537   convective scheme within the SPEEDY results. [*Kim et al.*, 2008] show that implementation of

538   CMT leads to enhanced off-equatorial precipitation (roughly 5°-15° off the equator) and

539   decreased precipitation between 5°S-5°N. The wind response to implementation of CMT is

540   increased 850 mb westerlies between 10°S-10°N and also easterlies poleward of the enhanced

541   precipitation. However, these results are only valid west of 150°W for the December - February

542   climatological forcing used for the [*Kim et al.*, 2008] example. A fundamental concept of CMT

543   is that upper atmosphere momentum is transported to the surface via downdrafts around

544   convection. In our example, the generally enhanced precipitation (Figure 2c) between 15°S-5°N

545   drags westerly momentum from upper branch of the Walker circulation to the surface (Figure

546   2a). At the same time, weaker precipitation for the coupled IO with respect to the decoupled IO

547   does the opposite leaving enhanced easterlies between 5°N-20°N and south of 15°S. Thus,

548   implementation of CMT within SPEEDY, but not within any of the linear model results, may

549   also contribute to the off-equatorial easterlies found in the SPEEDY results. Although there are

550   multiple potential reasons for differences between the nonlinear LBM and SPEEDY atmospheric

551   results, exploring differences further is beyond the scope of the current paper.

552   The combined impact of the zonal and meridional winds in the Pacific on the ocean can

553   be conveniently summarized by diagnosing the differences of the SPEEDY experiments using

554 curl and divergence of the wind stress. On the equator, surface convergence of the wind leads to

555 convergence of the surface currents, downwelling in the ocean, a deepening of the thermocline,

556 and an increase in sea level. The divergence of the atmospheric teleconnections is presented in

557 Figure 2f and this shows pervasive downwelling favorable winds all along the entire equator.

558 Off the equator, the curl can be used to estimate the sense of Ekman pumping velocity as a

559 measure of upwelling or downwelling. Figure 2e shows that upwelling favorable winds (positive

560 in the Northern Hemisphere and negative in the Southern Hemisphere) are predominant between

561 15°S to 10°N in the Pacific. However, the curl just to the south of the equator between 160°E-

562 140°W is positive indicating a narrow band of downwelling favorable curl. North of $\sim 10^{o}$N and

563 in the southwest Pacific (off Australia) downwelling curl is also prevalent.

564   To summarize differences between the present study and prior work, forced

565 experimental minus control atmospheric simulations produce strong easterly differences south of

566 15°S and between the equator and 20°N in the Pacific, with weak westerly differences near the

567 equator between 10°S and the equator. For the meridional component, winds converge towards

568 5°S especially over the eastern half of the basin with abundant precipitation in the eastern Pacific

569 cold tongue region near the equator. In addition, differences show strong positive precipitation

570 anomalies in the central Pacific. Convergence along the equator indicates that there is pervasive

571 downwelling favorable conditions present at initialization of the coupled system along the

572 equator. However, off the equator between 15°S-10°N the prevailing curl indicates that IO

573 SSTA is generally forcing upwelling in this region. As [*Annamalai et al.*, 2005] noted, the weak

574 winds within the waveguide may allow nascent El Niño/La Niña events to grow unencumbered.

575

6 Summary and Conclusions

Our results suggest that additional validated forecast skill is available for operational ENSO forecasting improvements by including IO forcing and realistically nonlinear modelling of the response of the coupled Indo-Pacific ocean atmosphere system. A key potential source that we have identified in this study is the impact of atmospheric teleconnections originating from the IO. Coupled experiments described herein that are initialized with the full observed SST forcing and utilize a nonlinear atmosphere indicate that our atmospheric response in the Pacific to interannual IO forcing includes weak westerly winds equatorward of 10^oS, enhanced off-equatorial trade winds and strengthened easterlies between 30^oS to 15^oS and the equator to 25^oN. The differences between the previous linear atmospheric model and our AGCM (SPEEDY) results may be due to either simplification of IO forcing, displacement of precipitation (and heating) to the east with no compensating anomaly, or most likely, the convective momentum transport in SPEEDY. These off-equatorial winds have profound impact in that they generate wind stress curl that act to amplify the oceanic Rossby wave signal which eventually impact the eastern Pacific by way of reflected Kelvin waves.

Differences between coupled experiments show that including the impact of interannual teleconnections from the IO have significantly higher ENSO correlation (exceeding the 95% significance level from 3-9 months) and lower RMS validation statistics. The reason for this is a combination of equatorial and off-equatorial coupling that eventually warms the NINO3 region. Early in the forecast period, prior to 3 month lead times, equatorial upwelling in the western Pacific weakly cools the NINO3 region via propagation of upwelling Kelvin waves. After that time, off-equatorial downwelling favorable curl in the western Pacific helps to amplify the transition from cooling to warming in the NINO3 region by way of reflected downwelling Rossby to downwelling Kelvin waves. Downwelling Kelvin waves, amplified by equatorial

600   convergence, warm the eastern Pacific and improve correlation validation after 3 month lead

601   times with respect to observations. The improvement in correlation peaks at 7 months which

602   corresponds with the time it takes for the transmission of the reflected downwelling Rossby wave

603   to reflect into the downwelling Kelvin wave then to propagate across the Pacific into the NINO3

604   region. Therefore, a main conclusion from these results is that the interannual variability of IO

605   SST forcing is responsible for overall somewhat lagged widespread downwelling in the Pacific,

606   assisted by off-equatorial curl, leading to warmer NINO3 SST anomaly and improved validation

607   after 3 month lead times.

608   　　　Currently ENSO forecast discussions (see http://origin.cpc.ncep.noaa.gov/

609   products/GODAS/ocean_briefing_gif/global_ocean_monitoring_current.ppt) include

610   descriptions of large-scale ocean waves present in the initialization of coupled forecasts.

611   However, these discussions only assess the state of the oceanic Kelvin wave (using the Ocean

612   Kelvin Wave Index, an extended EOF technique) and this would suggest a lack of emphasis on

613   off-equatorial processes in coupled model initialization. On the contrary, the results of the

614   impact of the teleconnections from the IO to the Pacific presented herein demonstrate the

615   significance of the off-equatorial processes that generate oceanic Rossby waves. Therefore, we

616   recommend that the impact of the Rossby waves on ENSO should be included in forecast

617   discussions. We have shown that the impact of the IO atmospheric teleconnections to ENSO

618   significantly improve coupled forecasts from 3-9 month lead times, so both upstream IO

619   influences and off-equatorial processes should be considered/included in ENSO forecasting

620   systems.

References

Adler, R. F., et al. (2003), The version-2 global precipitation climatology project (GPCP) monthly precipitation analysis (1979-present), *Journal of Hydrometeorology*, *4*(6), 1147-1167.

Alexander, M. A., I. Blade, M. Newman, J. R. Lanzante, N. C. Lau, and J. D. Scott (2002), The atmospheric bridge: The influence of ENSO teleconnections on air-sea interaction over the global oceans, *Journal of Climate*, *15*(16), 2205-2231.

Annamalai, H., S. Kida, and J. Hafner (2010), Potential Impact of the Tropical Indian Ocean-Indonesian Seas on El Nino Characteristics, *Journal of Climate*, *23*(14), 3933-3952.

Annamalai, H., S. P. Xie, J. P. McCreary, and R. Murtugudde (2005), Impact of Indian Ocean sea surface temperature on developing El Niño, *Journal of Climate*, *18*(2), 302-319.

Antonov, J. I., D. Seidov, T. P. Boyer, R. A. Locarnini, A. V. Mishonov, H. E. Garcia, O. K. Baranova, M. M. Zweng, and D. R. Johnson (2010), World Ocean Atlas 2009, Volume 2: Salinity., in *NOAA Atlas NESDIS 69*, edited by Secondary Antonov, J. I., D. Seidov, T. P. Boyer, R. A. Locarnini, A. V. Mishonov, H. E. Garcia, O. K. Baranova, M. M. Zweng, and D. R. Johnson, p. 184 pp., U.S. Government Printing Office, Washington, D.C.

Behringer, D. (2007), The Global Ocean Data Assimilation System (GODAS) at NCEP, paper presented at 11th Symp. on Integrated Observing and Assimilation Systems for Atmosphere, Oceans and Land Surface, American Meteorological Society San Antonio, TX [Available online at http://ams.confex.com/ams/pdfpapers/119541.pdf].

Bracco, A., F. Kucharski, F. Molteni, W. Hazeleger, and C. Severijns (2005), Internal and forced modes of variability in the Indian Ocean, *Geophysical Research Letters*, *32*(12).

Chen, D., L. M. Rothstein, and A. J. Busalacchi (1994), A Hybrid Vertical Mixing Scheme and Its Application to Tropical Ocean Models, *Journal of Physical Oceanography*, *24*(10), 2156-2179.

Delcroix, T., J. P. Boulanger, F. Masia, and C. Menkes (1994), Geosat-Derived Sea-Level and Surface Current Anomalies in the Equatorial Pacific During the 1986-1989 El-Niño and La-Niña, *Journal of Geophysical Research-Oceans*, *99*(C12), 25093-25107.

ECMWF (1994), The description of the ECMWF/WCRP Level III-A Global Atmospheric Data Archive, *Tech. Attach.Rep.*, 72 pp. pp, European Centre for Medium-Range Weather Forecasts, Reading, England, U.K.

Gent, P. R., and M. A. Cane (1989), A Reduced Gravity, Primitive Equation Model of the Upper Equatorial Ocean, *Journal of Computational Physics*, *81*(2), 444-480.

670   Gibson, J. K., P. Kallberg, S. Uppala, F. Hernandez, R. Nomura, and Y. Serrano (1997),
671   ECMWF re-analysis. *Rep.*, European Centre for Medium-Range Weather Forecasts,
672   Reading (UK).

673   Glantz, M. H. (2001), *Currents of Change: El Nino and La Nina Impacts on Climate and*
674   *Society*, 252 pp., Cambridge University Press, Cambridge, UK.

675   Griffies, S. M., M. J. Harrison, R. C. Pacanowski, and A. Rosati (2004), Techical Guide to
676   MOM4, in *GFDL Ocean Group Technical Report 5* edited by Secondary
677   Griffies, S. M., M. J. Harrison, R. C. Pacanowski, and A. Rosati, p. 337, GFDL GFDL
678   Princeton, NJ [Available online at www.gfdl.noaa.gov/~fms].

679   Hackert, E., A. Busalacchi, and J. Ballabrera-Poy (2014), Impact of Aquarius sea surface salinity
680   observations on coupled forecasts for the tropical Indo-Pacific Ocean, *Journal of*
681   *Geophysical Research, Oceans*, *119*(doi: 10.1002/2013JC009697).

682   Hackert, E., J. Ballabrera-Poy, A. J. Busalacchi, R. H. Zhang, and R. Murtugudde (2007),
683   Comparison between 1997 and 2002 El Nino events: Role of initial state versus forcing,
684   *Journal of Geophysical Research-Oceans*, *112*(C1).

685   Hackert, E. C., A. J. Busalacchi, and R. Murtugudde (2001), A wind comparison study using an
686   ocean general circulation model for the 1999-1998 El Niño, *Journal of Geophysical*
687   *Research-Oceans*, *106*(C2), 2345-2362.

688   Horel, J. D., and J. M. Wallace (1981), PLANETARY-SCALE ATMOSPHERIC
689   PHENOMENA ASSOCIATED WITH THE SOUTHERN OSCILLATION, *Monthly*
690   *Weather Review*, *109*(4), 813-829.

691   Huang, R. X. (1993), Real Fresh-Water Flux as a Natural Boundary-Condition for the Salinity
692   Balance and Thermohaline Circulation Forced by Evaporation and Precipitation, *Journal*
693   *of Physical Oceanography*, *23*(11), 2428-2446.

694   Kalnay, E., et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bulletin of the American*
695   *Meteorological Society*, *77*(3), 437-471.

696   Kessler, W. S., M. J. McPhaden, and K. M. Weickmann (1995), Forcing of Intraseasonal Kelvin
697   Waves in the Equatorial Pacific, *Journal of Geophysical Research-Oceans*, *100*(C6),
698   10613-10631.

699   Kim, D., J. S. Kug, I. S. Kang, F. F. Jin, and A. T. Wittenberg (2008), Tropical Pacific impacts
700   of convective momentum transport in the SNU coupled GCM, *Climate Dynamics*, *31*(2-
701   3), 213-226.

702   Kraus, E. B., and J. S. Turner (1967), A One-Dimensional Model of Seasonal Thermocline .2.
703   General Theory and Its Consequences, *Tellus*, *19*(1), 98-&.

704   Kroeger, J., and F. Kucharski (2011), Sensitivity of ENSO characteristics to a new interactive
705   flux correction scheme in a coupled GCM, *Climate Dynamics*, *36*(1-2), 119-137.

706 Kucharski, F., F. Molteni, and J. H. Yoo (2006), SST forcing of decadal Indian Monsoon rainfall
707   variability, *Geophysical Research Letters*, *33*(3).

708 Kucharski, F., I. S. Kang, R. Farneti, and L. Feudale (2011), Tropical Pacific response to 20th
709   century Atlantic warming, *Geophysical Research Letters*, *38*.

710 Kucharski, F., F. Molteni, M. P. King, R. Farneti, I.-S. Kang, and L. Feudale (2013), On the
711   Need of Intermediate Complexity General Circulation Models A "SPEEDY" Example,
712   *Bulletin of the American Meteorological Society*, *94*(1), 25-30.

713 Lau, N. C., and M. J. Nath (2003), Atmosphere-ocean variations in the Indo-Pacific sector during
714   ENSO episodes, *Journal of Climate*, *16*(1), 3-20.

715 Locarnini, R. A., A. V. Mishonov, J. I. Antonov, T. P. Boyer, H. E. Garcia, O. K. Baranova, M.
716   M. Zweng, and D. R. Johnson (2010), World Ocean Atlas 2009, Volume 1: Temperature,
717   NOAA Atlas NESDIS 68, edited by Secondary   Locarnini, R. A., A. V.
718   Mishonov, J. I. Antonov, T. P. Boyer, H. E. Garcia, O. K. Baranova, M. M. Zweng, and
719   D. R. Johnson, p. 184 pp., U.S. Government Printing Office, Washington, D.C.

720 McPhaden, M. J. (2015), Playing hide and seek with El Nino, *Nature Clim. Change*, *5*(9), 791-
721   795.

722 Menkes, C. E., M. Lengaigne, J. Vialard, M. Puy, P. Marchesiello, S. Cravatte, and G. Cambon
723   (2014), About the role of Westerly Wind Events in the possible development of an El
724   Nino in 2014, *Geophysical Research Letters*, *41*(18), 6476-6483.

725 Min, Q., J. Su, R. Zhang, and X. Rong (2015), What hindered the El Nino pattern in 2014?,
726   *Geophysical Research Letters*, *42*(16), 6762-6770.

727 Molteni, F. (2003), Atmospheric simulations using a GCM with simplified physical
728   parametrizations. I: model climatology and variability in multi-decadal experiments,
729   *Climate Dynamics*, *20*(2-3), 175-191.

730 Murtugudde, R., and A. J. Busalacchi (1998), Salinity effects in a tropical ocean model, *Journal
731   of Geophysical Research-Oceans*, *103*(C2), 3283-3300.

732 Murtugudde, R., R. Seager, and A. Busalacchi (1996), Simulation of the tropical oceans with an
733   ocean GCM coupled to an atmospheric mixed-layer model, *Journal of Climate*, *9*(8),
734   1795-1815.

735 Murtugudde, R., A. J. Busalacchi, and J. Beauchamp (1998), Seasonal-to-interannual effects of
736   the Indonesian Throughflow on the tropical Indo-Pacific Basin, *Journal of Geophysical
737   Research-Oceans*, *103*(C10), 21425-21441.

738 National Academies of Sciences, E., and Medicine (2016), *Next Generation Earth System
739   Prediction: Strategies for Subseasonal to Seasonal Forecasts*, 290 pp., The National
740   Academies Press, Washington, DC.

741 National Research Council (2010), *Assessment of intraseasonal to interannual climate prediction and predictability*, National Academies Press, Washington, D.C. :.

743 Picaut, J., and R. Tournier (1991), MONITORING THE 1979-1985 EQUATORIAL PACIFIC CURRENT TRANSPORTS WITH EXPENDABLE BATHYTHERMOGRAPH DATA, *Journal of Geophysical Research: Planets*, *96*, 3263-3277.

746 Price, J. F., R. A. Weller, and R. Pinkel (1986), Diurnal Cycling - Observations and Models of the Upper Ocean Response to Diurnal Heating, Cooling, and Wind Mixing, *Journal of Geophysical Research-Oceans*, *91*(C7), 8411-8427.

749 Quenouille, M. H. (1952), *Associated measurements*, 242 pp., Buterworths, London.

750 Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan (2003), Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century, *Journal of Geophysical Research-Atmospheres*, *108*(D14).

754 Reynolds, R. W., N. A. Rayner, T. M. Smith, D. C. Stokes, and W. Q. Wang (2002), An improved in situ and satellite SST analysis for climate, *Journal of Climate*, *15*(13), 1609-1625.

757 Rossow, W. B., and R. A. Schiffer (1991), Isccp Cloud Data Products, *Bulletin of the American Meteorological Society*, *72*(1), 2-20.

759 Saha, S., et al. (2014), The NCEP Climate Forecast System Version 2, *Journal of Climate*, *27*(6), 2185-2208.

761 Saha, S., et al. (2010), THE NCEP CLIMATE FORECAST SYSTEM REANALYSIS, *Bulletin of the American Meteorological Society*, *91*(8), 1015-1057.

763 Santoso, A., M. H. England, and W. Cai (2012), Impact of Indo-Pacific Feedback Interactions on ENSO Dynamics Diagnosed Using Ensemble Climate Simulations, *Journal of Climate*, *25*(21), 7743-7763.

766 Steiger, J. H. (1980), TESTS FOR COMPARING ELEMENTS OF A CORRELATION MATRIX, *Psychological Bulletin*, *87*(2), 245-251.

768 Uehara, H., A. A. Kruts, H. Mitsudera, T. Nakamura, Y. N. Volkov, and M. Wakatsuchi (2014), Remotely propagating salinity anomaly varies the source of North Pacific ventilation, *Progress in Oceanography*, *126*, 80-97.

771 Watanabe, M., and F. F. Jin (2003), A moist linear baroclinic model: Coupled dynamical-convective response to El Nino, *Journal of Climate*, *16*(8), 1121-1139.

773 Wu, R. G., and B. P. Kirtman (2004), Understanding the impacts of the Indian Ocean on ENSO variability in a coupled GCM, *Journal of Climate*, *17*(20), 4019-4031.

775 Xie, P. P., and P. A. Arkin (1998), Global monthly precipitation estimates from satellite-
776   observed outgoing longwave radiation, *Journal of Climate*, *11*(2), 137-164.

777 Yu, J. Y., C. R. Mechoso, J. C. McWilliams, and A. Arakawa (2002), Impacts of the Indian
778   Ocean on the ENSO cycle, *Geophysical Research Letters*, *29*(8).

779 Zhou, L., R. B. Neale, M. Jochum, and R. Murtugudde (2012), Improved Madden-Julian
780   Oscillations with Improved Physics: The Impact of Modified Convection
781   Parameterizations, *Journal of Climate*, *25*(4), 1116-1136.

782

783
784

Figure 1: Validation of the SPEEDY coupled model. Our Indo-Pacific SPEEDY coupled model (red) is a) correlated and b) RMSD against observed NINO3 SSTA for Jan. 1993-Mar. 2011. CFSRR coupled model results (black) are included to put our coupled results into the context of a more widely known coupled model. Individual correlations exceed the 95% significance out to 8.3 (35) and 10 months (34) (effective degrees of freedom) for red and black lines, respectively. Observation persistence is indicated by the thin black dotted line. This version of the SPEEDY coupled model assimilates all available satellite (sea level, SST) and in situ information (sea surface salinity and subsurface temperature and salinity) using the data assimilation technique described in [*Hackert et al.*, 2014].

Figure 2: Impact of interannual IO SST forcing. Differences between two sets of SPEEDY atmosphere-only experiments for a) zonal, b) meridional wind stress, c) precipitation, d) vector representation of a) and b), e) curl and f) divergence of the wind stress. Differences are full SST anomaly forcing over the Indo-Pacific region (i.e. INT_PAC, INT_IO) minus the experiment that uses climatological seasonal cycle forcing over the IO (INT_PAC, CLIM_IO). Letters "U" and "D" represent regions of upwelling and downwelling favorable winds and absolute values greater than 3.3×10^{-3} N/m^2, 2.7×10^{-3} N/m^2, 13.2 mm/mon, 0.53×10^{-9} N/m^3, 0.35×10^{-9} N/m^3 are significant at the 95% level for a), b) c), e), and f), respectively.

Figure 3: Impact of IO interannual forcing on coupled NINO3 SST results. Validation statistics for a) correlation and b) RMS differences between coupled experiments with full atmospheric coupling (i.e. INT_PAC, INT_IO) in red and interannual coupling in the Pacific and climatological forcing in the IO (i.e. INT_PAC, CLIM_IO) in blue. The coupled experiments are validated against observed NINO3 SST anomaly for 1993 to 2014. Individual correlations exceed the 95% significance out to 8 (43) and 5.8 months (41) (effective degrees of freedom) for red and blue lines, respectively. The thick black line on the top x-axis shows where the red line is significantly larger than the blue line using the Steiger-Z test.

Figure 4: Mean and variability of NINO3 SST for impact of interannual IO SST forcing. Plots showing the NINO3 SST a) mean forecast and b) variability for INT_PAC, INT_IO (red) and INT_PAC, CLIM_IO (blue) for all forecasts from 1993-2014.

Figure 5: Hovmöller plots of impact of interannual IO SST forcing. Plots showing the mean temporal evolution of the impact of IO atmospheric coupling using longitude versus lead time (in months) averaged between 2°N and 2°S for a) SST, b) sea level (SL), c) zonal wind stress, and d) zonal currents. The mean is taken for the average forecast differences, INT_PAC, INT_IO minus INT_PAC, CLIM_IO, over all months from 1993 to 2014.

Figure 6: Kelvin/Rossby wave decomposition of interannual IO SST forcing. Longitude versus time distribution of the equatorial (a) Kelvin and (b) the first meridional mode of equatorial Rossby waves through their signature in zonal surface current deduced from the average forecast SL differences, (INT_PAC, INT_IO) – (INT_PAC, CLIM_IO). In order to follow possible wave reflections on the western (WB) and eastern (EB) boundaries, the Rossby panel (b) is inverted and the Kelvin wave pattern is repeated (c). The color scale for the Rossby panel is also inverted since reflection on meridional boundaries results in zonal currents of opposite sign. Solid lines (downwelling) and dashed lines (upwelling) represent theoretical wave speeds for Kelvin (2.5m/s) and Rossby waves (-0.8m/s or ~5months to cross this Pacific basin at 5°N) on each plot.

827   **Figure 7:** Average 3 month forecast INT_PAC, INT_IO - INT_PAC, CLIM_IO. Average
828   forecast values for month 3 for a) SST, b) sea level, c) curl of the wind stress (color) and wind
829   stress (vector), and d) divergence of the wind stress (color) and wind stress (vector). The scale
830   of the vector plot is indicated in the bottom left of the panel. For the reader's convenience,
831   regions of upwelling and downwelling are marked by letters U and D, respectively.

832   **Figure 8:** Average 5 month forecast INT_PAC, INT_IO - INT_PAC, CLIM_IO. Same as
833   previous but for 5 month average forecasts.

834   **Figure 9:** Average 7 month forecast INT_PAC, INT_IO - INT_PAC, CLIM_IO. Same as
835   previous but for 7 month lead forecast mean.

836   **Figure 10:** Average 10 month forecast INT_PAC, INT_IO - INT_PAC, CLIM_IO. Same as
837   previous but for 10 month forecast mean.
838

Experiment Name	Period	Model Geometry (Atmosphere/Ocean)	Indo-Pacific Forcing
INT_PAC, INT_IO	1993-2014	Global/Indo-Pacific	Interannual SSTA forcing for Pacific and IO
INT_PAC, CLIM_IO	1993-2014	Global/Indo-Pacific	Interannual SSTA forcing for Pacific, climatological seasonal cycle SST for IO

Table 1: Experiment description for impact of interannual IO SST forcing. The far left column describes the experiments, "INT" and "CLIM" stand for interannual and climatological forcing and "PAC" and "IO" stand for Pacific (30°S-30°N, 130°E-70°W) and Indian Oceans (30°S-30°N, 30°E-129°E), respectively. The far right column describes the SST anomaly forcing (SSTA) for the Indo-Pacific region. In order to isolate the impact of the IO, differences between INT_PAC, INT_IO – INT_PAC, CLIM_IO are presented. Note that SSTA are formulated with respect to the 1983-2014 mean seasonal cycle using [Reynolds et al., 2002] OI SST.

Figure 1: Validation of the SPEEDY coupled model. Our Indo-Pacific SPEEDY coupled model (red) is a) correlated and b) RMSD against observed NINO3 SSTA for Jan. 1993-Mar. 2011. CFSRR coupled model results (black) are included to put our coupled results into the context of a more widely known coupled model. Individual correlations exceed the 95% significance out to 8.3 (35) and 10 months (34) (effective degrees of freedom) for red and black lines, respectively. Observation persistence is indicated by the thin black dotted line. This version of the SPEEDY coupled model assimilates all available satellite (sea level, SST) and in situ information (sea surface salinity and subsurface temperature and salinity) using the data assimilation technique described in [*Hackert et al.*, 2014].

Figure 2: Impact of interannual IO SST forcing. Differences between two sets of SPEEDY atmospheric-only experiments for a) zonal, b) meridional wind stress, c) precipitation, d) vector representation of a) and b), e) curl and f) divergence of the wind stress. Differences are full SST anomaly forcing over the Indo-Pacific region (i.e. INT_PAC, INT_IO) minus the experiment that uses climatological seasonal cycle forcing over the IO (INT_PAC, CLIM_IO). Letters "U" and "D" represent regions of upwelling and downwelling favorable winds and absolute values greater than 3.3×10^{-3} N/m^2, 2.7×10^{-3} N/m^2, 13.2 mm/mon, 0.53×10^{-9} N/m^3, 0.35×10^{-9} N/m^3 are significant at the 95% level for a), b) c), e), and f), respectively.

Figure 3: Impact of IO interannual forcing on coupled NINO3 SST results. Validation statistics for a) correlation and b) RMS differences between coupled experiments with full atmospheric coupling (i.e. INT_PAC, INT_IO in red and interannual coupling in the Pacific and climatological forcing in the IO (i.e. INT_PAC, CLIM_IO) in blue. The coupled experiments are validated against observed NINO3 SST anomaly for 1993 to 2014. Individual correlations exceed the 95% significance out to 8 (43) and 5.8 months (41) (effective degrees of freedom) for red and blue lines, respectively. The thick black line on the top x-axis shows where the red line is significantly larger than the blue line using the Steiger-Z test.

Figure 4: Mean and variability of NINO3 SST for impact of interannual IO SST forcing. Plots showing the NINO3 SST a) mean forecast and b) variability for INT_PAC, INT_IO (red) and INT_PAC, CLIM_IO (blue) for all forecasts from 1993-2014.

Figure 5: Hovmöller plots of impact of interannual IO SST forcing. Plots showing the mean temporal evolution of the impact of IO atmospheric coupling using longitude versus lead time (in months) averaged between 2°N and 2°S for a) SST, b) sea level (SL), c) zonal wind stress, and d) zonal currents. The mean is taken for the average forecast differences, INT_PAC, INT_IO minus INT_PAC, CLIM_IO, over all months from 1993 to 2014.

Figure 6: Kelvin/Rossby wave decomposition of interannual IO SST forcing. Longitude versus time distribution of the equatorial (a) Kelvin and (b) the first meridional mode of equatorial Rossby waves through their signature in zonal surface current deduced from the average forecast SL differences, (INT_PAC, INT_IO) – (INT_PAC, CLIM_IO). In order to follow possible wave reflections on the western (WB) and eastern (EB) boundaries, the Rossby panel (b) is inverted and the Kelvin wave pattern is repeated (c). The color scale for the Rossby panel is also inverted since reflection on meridional boundaries results in zonal currents of opposite sign. Solid lines (downwelling) and dashed lines (upwelling) represent theoretical wave speeds for Kelvin (2.5m/s) and Rossby waves (-0.8m/s or ~5months to cross this Pacific basin at 5°N) on each plot.
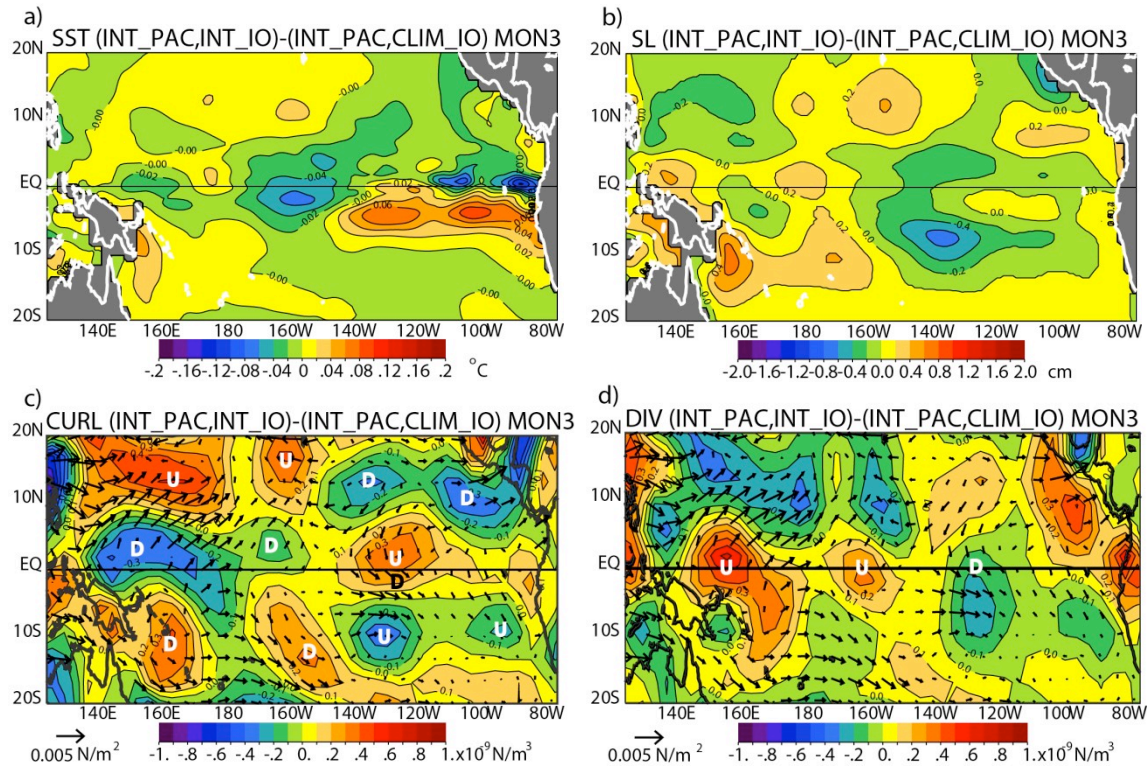
Figure 7: Average 3 month forecast INT_PAC, INT_IO - INT_PAC, CLIM_IO. Average forecast values for month 3 for a) SST, b) sea level, c) curl of the wind stress (color) and wind stress (vector), and d) divergence of the wind stress (color) and wind stress (vector). The scale of the vector plot is indicated in the bottom left of the panel. For the reader's convenience, regions of upwelling and downwelling are marked by letters U and D, respectively.
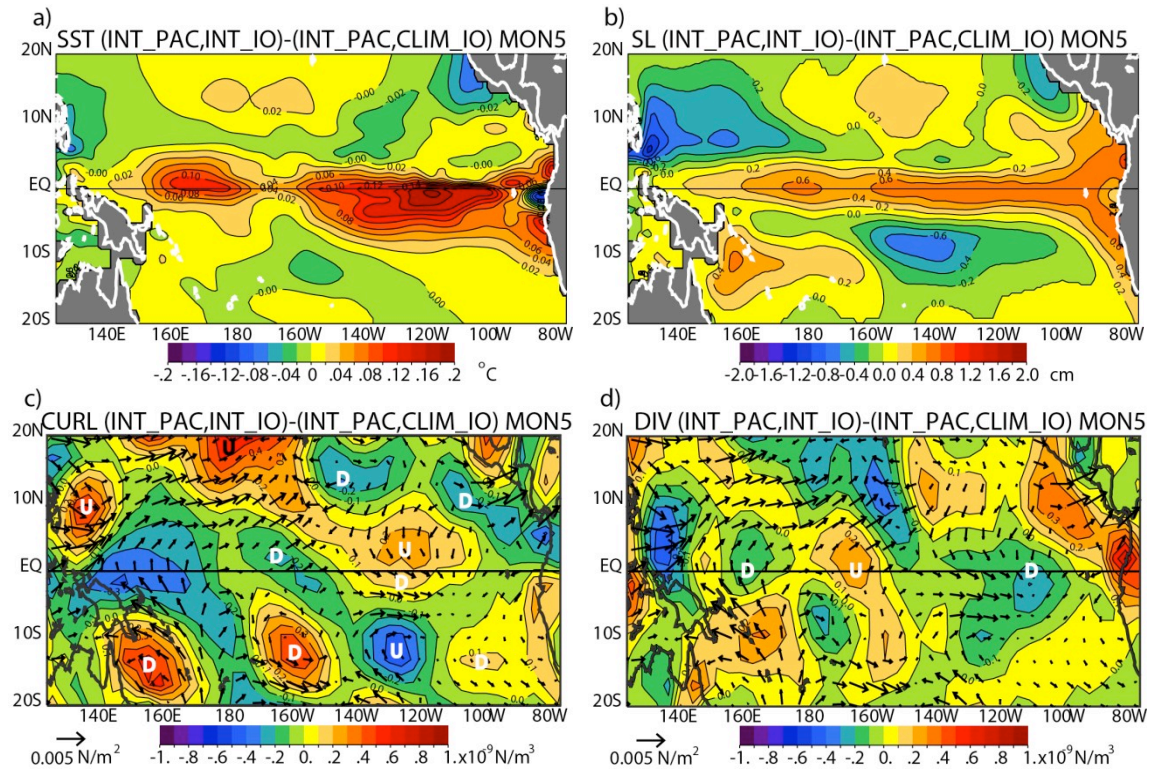
Figure 8: Average 5 month forecast INT_PAC, INT_IO - INT_PAC, CLIM_IO. Same as previous but for 5 month average forecasts.
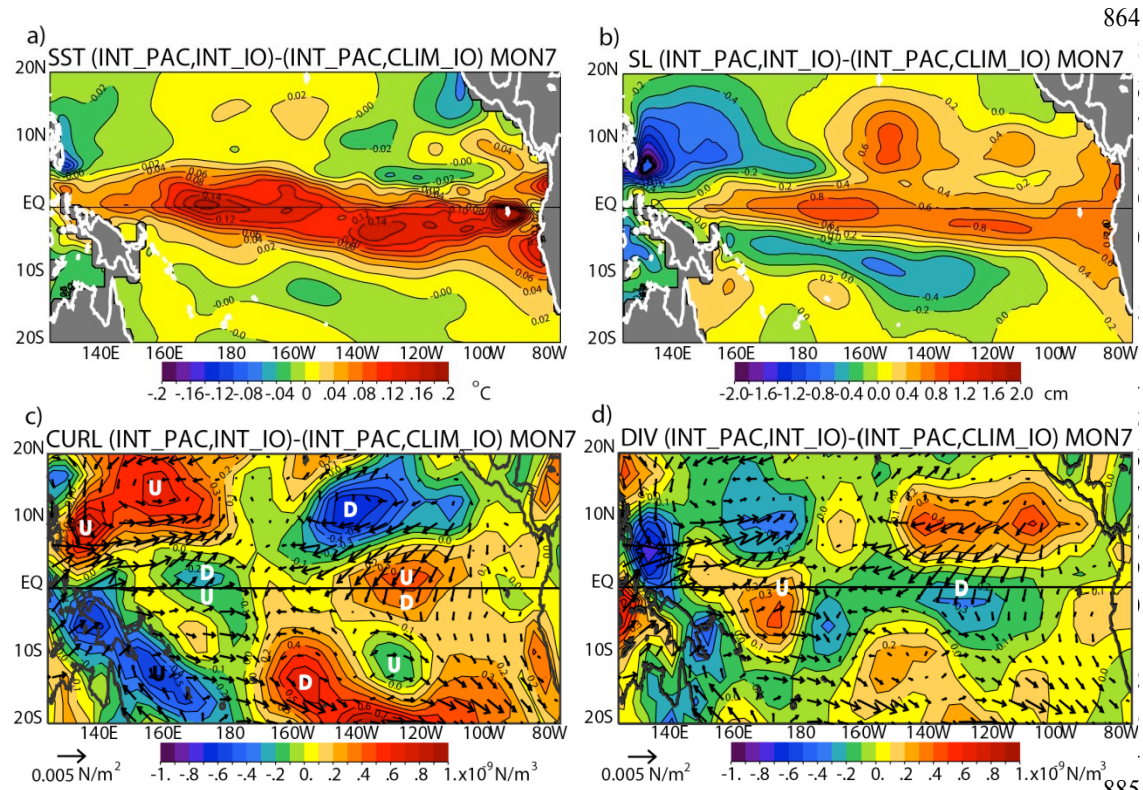
Figure 9: Average 7 month forecast INT_PAC, INT_IO - INT_PAC, CLIM_IO. Same as previous but for 7 month lead forecast mean.
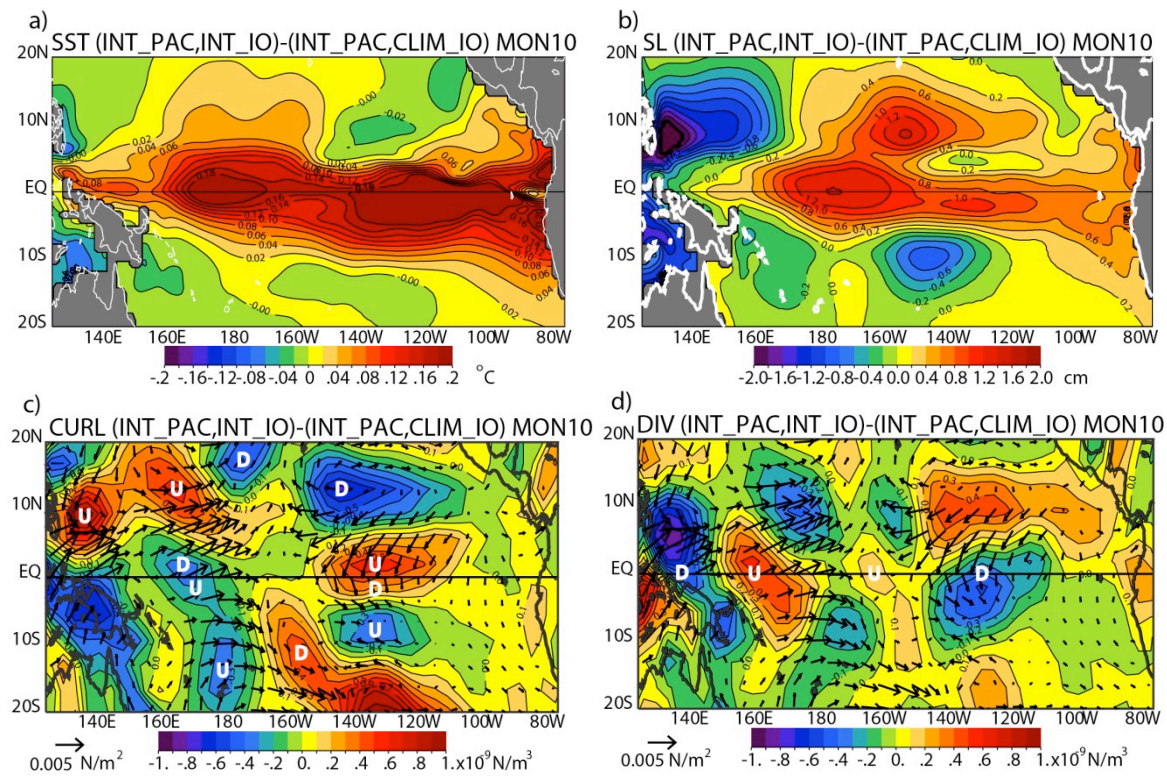
Figure 10: Average 10 month forecast INT_PAC, INT_IO - INT_PAC, CLIM_IO. Same as previous but for 10 month forecast mean.